

What are Evapotranspiration and Forecast Reference Crop Evapotranspiration (FRET)?

The purpose of this document is not to explain ET to the scientific community (they can turn to the references), but rather to provide the scientific context for many of the decisions, formulations, data sources, etc. related to the production of National Weather Service FRET products.

1. Introduction

Evapotranspiration *ET* is the sum of two similar processes: the evaporation of moisture from soil, water bodies, wet vegetated canopies and the transpiration of moisture drawn through and from within plants. Forecasts of this crucial link in the hydrologic cycle are particularly useful for water managers and agricultural users in the western US, where, due to widespread aridity and the otherwise spatially heterogeneous nature of water resources, water availability is at a premium.

Briefly, the *ET* flux is capped by the ability of the atmosphere to absorb and bear away moisture (the “drying power of the air”) and limited by the availability of two things: energy to drive the evaporative process (changing water from liquid to vapor) and water to evaporate and/or transpire. From observations of various meteorological factors—shortwave radiation (sunshine), longwave radiation, temperature, humidity, and wind speed—we can estimate the energy available for evaporation and the drying power of the air. There are many scientific formulations of *ET* that combine some or all of these drivers with knowledge of the energy limitation, resulting in *ET* estimates of widely ranging hydrologic merit. However, in the context of providing *ET* estimates to water managers and agricultural users, it is important to remember that all of these estimates depend on knowledge of the availability of moisture (the third limit mentioned above), and therein lies the rub: at the spatial scales these end-users require neither the quantity nor the spatial distribution of moisture at the land surface are knowable. Thus, it is almost impossible to *directly* estimate or observe *ET* usefully for water managers and agricultural users.

Thus, in estimating *ET*, while we seek to answer the fundamental question: “how much water is supplied from the earth’s surface to the atmosphere?”, we are instead forced to ask a simpler question: “given well-defined, ideal surface conditions, what could the land surface supply to the atmosphere?” An estimate of this second flux would then define a specific, or ideal, supply. The formalized answer to this question yields reference crop *ET* (ET_r) so called as the surface is conceived of as a specific crop. ET_r can then be used as a starting estimate from which additional assumptions as to prevailing soil- and vegetation-moisture conditions and the vegetation mix and phenology are applied to account for the divergence from the ideal, well-defined conditions to scale down for *ET*.

There are many anticipated users of such a forecast of evaporative demand, including individual farmers, municipal suppliers, federal and state agencies, and academia:

- Agricultural users, particularly irrigators, have frequently asked WFOs to provide *ET* grids. These users are likely to want a standardized ET_r estimate

- Water resource managers of sophisticated supply systems (e.g., Denver Water) may have an interest in individual inputs to the NWSRFS such as ET , particularly at a multi-week or seasonal time-scale as they forecast demand and arrange for trans-basin supplies to meet the demand. They might prefer forecasts with lead-times of months or seasons, time-frames at which even simply getting the sign (i.e., the direction: up or down) of a change right would be advantageous.
- The Bureau of Reclamation (USBR), for planning reservoir releases, especially a week or two out.
- The US Forest Service for water supply.
- The academic / agricultural outreach community, particularly from UC Davis, whose faculty has been assisting the National Weather Service Sacramento Forecast Office in this effort.
- The US Geological Survey (USGS) Earth Resources Observation and Science (EROS) division has also expressed an interest in becoming a dedicated user of the climatological dataset that underpins/adds value to the forecast. They would use these data to assist in their estimation of actual ET using remote sensing and energy balance principles. [Gabriel Senay, USGS, personal communication, 2010].
- The National Integrated Drought Information System (NIDIS) is seeking a dataset of evaporative demand (such as ET_n) that has a short latency and is consistent with a historical period. Such a dataset would be highly desirable for drought analysis, allowing for rapid analysis of real-time changes in drought conditions. Forecasts of evaporative demand would be “icing on the cake,” permitting an expectation of near-future changes in drought conditions [Jim Verdin, NIDIS, personal communication, 2010].

2. Estimating reference crop evapotranspiration ET_{rc}

2.1 Overview

Ultimately, ET is the flux of interest, as it is the actual flux of moisture from the surface to the atmosphere under prevailing conditions. A physically sound measure of ET would convert net available energy Q_n input to moisture flux through a function of wind speed U_z , temperature T , humidity e_a , and the availability of moisture to evaporate Θ .

$$ET = f(Q_n, U_z, T, e_a, \Theta) \quad (1)$$

However, due to the lack of knowledge about the state of Θ on useful spatial and temporal scales, on an operational basis ET cannot be estimated directly using all appropriate drivers (Equation 1). The most common ET observation technique is based on water balances across defined basins, wherein the ET rate is estimated as a residual of a water balance that accounts for all other moisture fluxes into and from, and storage changes within, the basin. In the forecast context, the utility of such techniques is limited to calibrating other operational ET formulations.

More often, the difficulties pertaining to Θ are overcome by estimating ET_{rc} , which is an idealized supply of moisture from the land surface to the atmosphere. Physically based formulations of ET_{rc} are attempts to model both the radiative (or surface energy) and

advective drivers and, similarly to physically sound ET formulations, they convert Q_n input to moisture flux as a function of U_z , T , e_a :

$$ET_{rc} = f(Q_n, U_z, T, e_a) \quad (2)$$

Most formulations derive ET by scaling ET_{rc} down by a series of coefficients k_i that describe various field conditions (e.g., single or mixed cropping, soil-water stress, salinity), thereby yielding ET values between zero for dry conditions and ET_{rc} for unlimited moisture and ideal conditions:

$$ET = f\{ET_{rc}, k_1, k_2, k_3, \dots\} \quad (3)$$

The Penman-Monteith formulation (see Equation 12) estimates ET directly and is soundly physically based: it obviates direct knowledge about Θ by using as surrogates a series of variable resistances to the diffusion of water vapor from within the leaves of a vegetative canopy to the overpassing air. In the Penman-Monteith framework, ET_{rc} is a specific case of ET with very strictly prescribed surface conditions: a hypothetical, well-watered crop of height b of 0.12 m, a stomatal resistance r_s of 70 sec/m, and an albedo α of 0.23. ET_{rc} so defined is supposed to mimic the evaporation from an extensive surface of adequately watered, actively growing green grass of uniform height.

The surface energy balances (i.e., Q_n) of the two formulations used to provide FRET forecasts and climatologies are similar, and these are addressed in Section 2.2. More obvious differences between the two formulations arise in their advective components: these are then addressed as part of each formulation's description in Sections 2.3.1 and 2.3.2. There are also simpler formulations of ET_{rc} that are based only on temperature, but these less physically sound formulations do not explicitly attempt to model the surface energy balance and so make no assumptions about the nature of the evaporating surface. They are not used to provide FRET forecasts or climatologies, and so are not discussed further.

2.2 Estimating the radiative driver/surface energy balance

The energy input to an evaporating surface to drive ET is provided by the absorption at the surface of solar (shortwave) radiation and thermal (longwave) radiation. This absorption is the incidence at the surface of the radiation fluxes less (i) the reflection of shortwave and (ii) the re-emission of longwave radiation, and this results in the net radiative balance. The energy available for evaporation at the surface Q_n is then the net radiative balance less the far-smaller fluxes to the ground and heat storage changes in the evaporating surface, and the relative scales of these fluxes explains why herein Q_n is referred to as the “radiative driver.”

It is crucial, therefore, to have first an accurate estimate of the net radiative balance at the surface. However, complete knowledge of the net radiative balance requires extensive instrumentation not generally available on an operational basis. As a result, much parameterization is required, and generally this relies on, or is mixed with, observations of surrogate variables.

Toward the final aim of estimating the net available energy for evaporation Q_n for ET_{rc} , the following section describes the surface energy balance of an evaporating surface. The

treatments of each component of Q_n is then described as they contribute (or do not contribute) to its derivation.

2.2.1 Surface energy balance

The instantaneous energy balance at an evaporating surface can be expressed by balancing the time-rate of change of heat-storage in the evaporating layer as the residual of all fluxes into and from the layer, as shown in Figure 1 below, and Equation (4) as follows:

$$\frac{\partial W}{\partial t} = (1 - \alpha)R_{in} + L_{in} - L_{out} - H - ET - G - C - A_d, \quad (4)$$

where all fluxes (RHS) and time-rates of change of heat storage (LHS) are in flux units [W/m^2], with positive fluxes into the evaporating surface, negative out. $\partial W/\partial t$ is the time-rate of change of heat storage in the evaporating layer (positive increase), α is the surface albedo [-], R_{in} is the downward shortwave radiation incident at the surface, L_{in} and L_{out} are respectively the longwave radiation fluxes inward to and outward from the surface, H is the sensible heat flux by diffusion from the surface, ET is the latent heat flux equivalent of actual evapotranspiration, G is the heat flux conducted into the soil (or ground heat flux) from below the evaporating layer, C is the energy absorbed by biochemical processes in vegetation in the control volume, and A_d is heat lost to advection from the control volume.

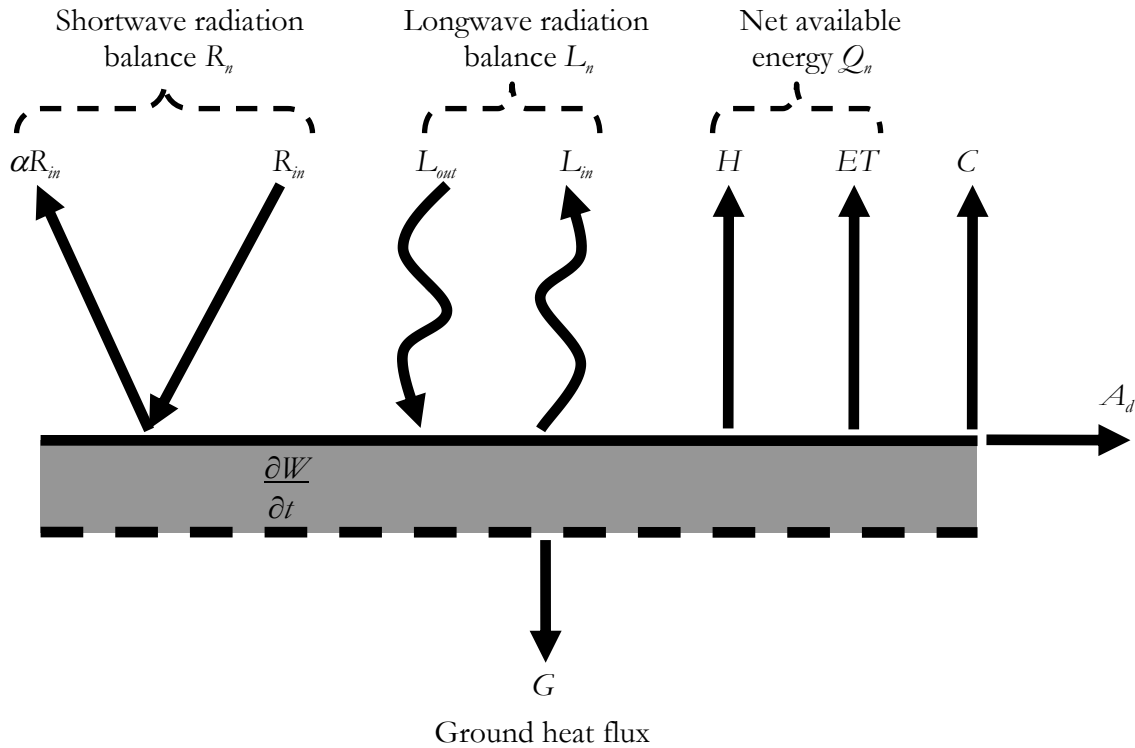


Figure 1: Instantaneous energy balance at an evaporating surface. All fluxes and heat-storage changes are in flux units [e.g., W/m^2]. The grey rectangle represents the control volume to and from which all fluxes pass and within which all heat-storage changes are considered, defined as an extensive free water surface for E_p , an evaporation pan filled with water for E_{pan} , and a ground surface and vegetated canopy for ET and ET_r .

The terms in the balance above are often conflated into groups that are more conceptually malleable. The radiation terms are gathered into R_n and L_n —respectively the net shortwave and longwave radiative fluxes to the surface—as follows:

$$R_n = (1 - \alpha)R_{in}, \quad (5)$$

and

$$L_n = L_{in} - L_{out}. \quad (6)$$

The net radiative balance (i.e., $R_n + L_n$) at the surface is then the sum of the differences between incoming and outgoing shortwave radiation and between incoming and outgoing longwave radiation. Generally, shortwave radiation is more often observed than is longwave radiation, but even then mostly as a function of such simple measures as sky cover (e.g., Equation 9) or sunshine hours; such data are widely available in time and space but suffer from issues of physical relevance. Most conveniently, the terms for net radiation balance R_n and L_n , conduction G through and heat storage changes $\partial W / \partial t$ within the evaporating layer, and the smaller heat fluxes associated with biochemistry C and advection A_d from the control volume are conceived of as together (LHS in Equation (7) below) providing the net available energy Q_n . Q_n is then partitioned into the processes of latent heat flux due to evapotranspiration ET and sensible heat flux H (RHS in Equation (7) below), as follows:

$$R_n + L_n - G - \frac{\partial W}{\partial t} - A_d - C = Q_n = ET + H. \quad (7)$$

2.2.2 Shortwave radiation balance

Parameterizations of R_n at the evaporating surface generally proceed from knowledge of the incoming shortwave radiation to the top of the atmosphere R_{toa} , and then account first for scattering, absorption and reflection of this flux as a function of dust, aerosols, clouds, and humidity through which it has to pass, and then the its reflection from the surface.

The extra-terrestrial or top-of-atmosphere shortwave radiation R_{toa} [W/m^2] is calculated from:

$$R_{toa} = 15.392 \frac{\lambda}{86400} d_r (\varpi_s \sin \phi \sin \delta + \cos \phi \cos \delta \sin \omega_s) \quad (8)$$

where the constant 15.392 represents the solar constant expressed as an evaporative equivalent [mm/day], ϕ is the latitude [rads, +ve N], and λ is the latent heat of vaporization [J/kg] [Shuttleworth, 1992]. The other variables are time-dependent: d_r is the relative distance from the earth to the sun on the J th day of the year [-], δ is the solar declination [rads], and ω_s is the sunset hour angle [rads]. R_{in} incident at the evaporating surface [W/m^2] is scaled down from R_{toa} as a function of either the ratio of observed to maximum sunlight hours n_i/N , or the percent cloud cover.

The traditional method of converting from R_{toa} to R_{in} is using the Savinov-Angström equation:

$$R_{in} = R_{toa} \left(1 - a_{sa} \frac{CC_{daily}}{100} \right), \quad (9)$$

where a_{sa} is a calibrated constant [-] of the order 0.71. CC_{daily} is the mean cloud cover [%] during daylight hours [Brutsaert, 1982].

Having estimated (or measured) R_{in} , R_n [W/m²] is then that portion of R_{in} that is absorbed by the evaporating surface, or simply R_{in} less the portion that is reflected, which is R_{in} multiplied by the dimensionless surface albedo α , as shown in Equation (5) above.

2.2.3 Longwave radiation balance

As there are few, if any, operational measurements of surface longwave radiation or its surrogates, the longwave radiation balance L_n is, in general, even more heavily parameterized than R_n . Both L_{in} and L_{out} are calculated according to the Stefan-Boltzmann Law, which states that a body's longwave emission L [W/m²] varies as a function of the fourth power of its absolute temperature T [K], or:

$$L = \sigma \epsilon T^4, \quad (10)$$

where σ is the Stefan-Boltzmann constant (5.6704×10^{-8} W/m²/K⁴), ϵ is the emissivity, which is an intrinsic property of the evaporating surface that quantifies its similarity of the evaporating surface to a purely emitting perfect black body (i.e., one for which $\epsilon = 1$).

However, various data constraints limit estimation of temperatures and emissivities. The only temperature datum known (i.e., T at 2 m) is used for both T and T_s , either from daily maximum and minimum temperatures T_{max} [K] and T_{min} [K], respectively, or from regular, intra-daily observations of T . With regard to the emissivities, L_{in} varies with atmospheric composition, particularly with such factors as clouds, dust, and concentrations of greenhouse gasses (e.g., water vapor and carbon dioxide). Each of these constituents has its own emissivity and is of unknown concentration so, to simplify, atmospheric concentrations are assumed constant. Then, given that the temperatures for both emitting bodies (i.e., the surface and the atmosphere) are estimated identically, their individual emissivities are wrapped up into a bulk or effective net emissivity ϵ' , parameterized as a function of surface humidity and of the clarity of the sky to shortwave radiation (i.e., the ratio R_{in}/R_{cs}).

The surface gains longwave energy under cloudy conditions, but loses longwave energy under clearer skies. Also note that, all else equal, positive and negative longwave exchanges at the surface vary with the square root of humidity, as measured by e_a .

2.2.4 Ground heat flux

Estimation of the ground heat flux G [W/m²] requires knowledge of soil temperature and moisture content and so is generally avoided, under the assumption that G over small time periods, such as low multiples of days, is generally orders of magnitude lower than the net radiative balance; thus G is often approximated to zero (also see *Allen et al.* [1998]).

2.2.5 Time-rate of change of heat storage

Over periods of days, the time-rate of change of heat storage $\partial W / \partial t$ is generally held to be negligible compared to the radiation balances, and so is set to zero.

2.2.6 Energy advected from the control volume

The energy flux advected from the control volume A_d may be transported by air in the case of ET and ET_{rc} or may include larger fluxes due to flows into and out of water bodies and precipitation onto water bodies in the case of E_p . This is neglected.

2.2.7 Biochemical/biomass-related heat storage

The uptake of heat energy in biochemical processes and biomass in the control volume C is generally negligible compared to the radiative and latent heat fluxes of interest—on the order of 2% of Q_n [Shuttleworth 1992]—and so is neglected.

2.2.8 Net available energy

As a starting point for formulation of Q_n , recall the generic expression for net available energy for evaporation Q_n from Equation (7), repeated here:

$$Q_n = R_n + L_n - G - \frac{\partial W}{\partial t} - A_d - C, \quad (7 \text{ again})$$

where all fluxes are in W/m^2 (multiply by 86400/ λ for mm/day). In this case, Q_n is defined as:

$$Q_n = R_n + L_n. \quad (11)$$

2.3 Formulations of reference crop evapotranspiration ET_{rc}

In this section the specifics of each of the physically based formulations of the evaporative fluxes of interest are detailed, paying particular reference to their advective drivers, wherein the primary differences arise.

2.3.1 Penman-Monteith ET_{rc}

As stated before, in the Penman-Monteith framework, ET_{rc} is merely a heavily specified case of the more-general formulation for ET . The Penman-Monteith formulation for ET is well documented and has gained national and international acceptance (see the FAO 56 report of Allen *et al.* [1998]). It is formulated to calculate ET by including in its advective conception a parameterization of the fine-scale diffusive characteristics of the plants and the surface under variable moisture availability conditions. Assuming that the onerous data and parameter requirements can be met, this versatile equation can then also be used to calculate the more specific cases of E_p and reference crop evaporation ET_{rc} , as described in following sections. The general formulation for ET is as follows:

$$ET = \frac{\Delta}{\Delta + \gamma \left(1 + \frac{r_s}{r_a}\right)} Q_n + \frac{\frac{\rho_a c_p}{r_a}}{\Delta + \gamma \left(1 + \frac{r_s}{r_a}\right)} (e_{sat} - e_a), \quad (12)$$

where ET is in W/m^2 (multiply by $86400/\lambda$ for mm/day), Q_n is in W/m^2 . Δ and γ are in Pa/K, respectively, e_{sat} and e_a are in Pa, ρ_a is the density [kg/m^3] of moist air, and c_p [$J/kg/K$] is as previously defined [Allen *et al.*, 1998]. R_s is the “stomatal resistance,” a bulk measure of the resistance [sec/m] to the diffusion of water vapor from stomata within the vegetative canopy, and r_a is the aerodynamic resistance [sec/m] to the diffusion of both water vapor and sensible heat from the canopy to the measurement height. R_a depends on land cover (e.g., crop height) while r_s reflects moisture availability restrictions, increasing under drier conditions.

According to Brutsaert [1982], use of the Penman-Monteith formulation for ET (Equation 12) is constrained primarily by the lack of knowledge about the status of r_s , more specifically the unknown distribution of heating by radiation within the canopy, the spatial variability of vapor sources on diurnal and seasonal cycles, seasonality of plant physiology, moisture stress at the roots, and species-specific physiology.

For the general case of ET -estimation, the parameter r_a may be estimated from:

$$r_a = \frac{\ln\left(\frac{z_m - d}{z_{0m}}\right) \ln\left(\frac{z_v - d}{z_{0v}}\right)}{k^2 U_z}, \quad (13)$$

with all other variables previously defined [Allen *et al.*, 1998]. For many crops, d , z_{0m} , and z_{0v} can be related to crop height h [m] as follows [Shuttleworth, 1992]:

$$d = \frac{2h}{3} \quad (14)$$

$$z_{0m} = 0.123h \quad (15)$$

$$z_{0v} = 0.1z_{0m}. \quad (16)$$

The familiar FAO-56 ET_r is merely an ET rate calculated for biological and physical conditions that are defined as follows: a hypothetical, well-watered crop of height h of 0.12 m, r_s of 70 sec/m, α of 0.23, and specifying that measurements of U_z , T , and e_a are taken at 2 m above the ground. Using the above approximations (Equations 14, 15, and 16) for d , z_{0m} , and z_{0v} yields r_a implicitly specified as $208/U_2$ [sec/m]. Using Q_n for the surface from Equation (11) yields the following expression for the Penman-Monteith formulation of ET_r :

$$ET_r = \frac{\Delta}{\Delta + \gamma(1 + 0.34U_2)} Q_n + \frac{\gamma}{\Delta + \gamma(1 + 0.34U_2)} \frac{0.9}{T} \frac{\lambda}{86400} U_2 (e_{sat} - e_a), \quad (17)$$

for ET_r in W/m^2 (multiply by $86400/\lambda$ for mm/day) [Shuttleworth, 1992]. All other variables are defined above for ET , with T [K], U_2 [m/sec], and e_a [Pa] specified at a 2-m height.

2.3.2 Kimberly Penman ET_r

The Kimberly version of the Penman combination equation for ET_r is conceptually similar to the Penman-Monteith expression for E_p . The primary differences are that, at time-scales longer than monthly, the net available energy Q_n for ET_r includes the effects of losses to ground heat flux G , and the parameters of the vapor transfer function in the Kimberly Penman equation are seasonalized as explicit functions of the Julian day of the year J [-]. The

data for the empirical calibration of the vapor transfer function come from Kimberly, ID, and the formulation is as follows:

$$ET_{rc} = \frac{\Delta}{\Delta + \gamma} Q_n + \frac{\gamma}{\Delta + \gamma} \frac{6.43}{86.4} (a_{KP} + b_{KP} U_2) (e_{sat} - e_a) \quad (18)$$

where ET_{rc} is in W/m^2 (multiply by 86400/ λ for mm/day) and Δ , γ , U_2 , e_{sat} , e_a are as previously defined. The 86.4 denominator accounts for a conversion of the aerodynamic portion (the 2nd term on the RHS) from mm/day and e_{sat} and e_a originally in kPa. The seasonal vapor transfer function parameters a_{KP} and b_{KP} are defined as follows (for the northern hemisphere) [Shuttleworth, 1992]:

$$a_{KP} = 0.4 + 1.4 \exp\left(-\left(\frac{J - 173}{58}\right)^2\right), \quad (19)$$

and

$$b_{KP} = 0.605 + 0.345 \exp\left(-\left(\frac{J - 243}{80}\right)^2\right). \quad (20)$$

2.4. Why forecast both Penman-Monteith and Kimberly Penman ET_{rc} ?

NWS Offices outside the Pacific Northwest, such as the Sacramento WFO (North Central California) and Hanford WFO (South Central California) calculate forecasts of Penman-Monteith ET_{rc} as adopted by the Environmental Water Resources Institute – American Society of Civil Engineers [Walter *et al.*, 2005], in collaboration with the University of California Davis and California Department of Water Resources.

NWS Offices in the Pacific Northwest, such as Pendleton WFO (northeast Oregon and southeast Washington) and Great Falls WFO (central Montana) currently use the Kimberly Penman equation (Equation 18) to predict (and publish) its ET_{rc} estimates. The rationale for this choice is that the parameters a_{KP} (Equation 19) and b_{KP} (Equation 20) were calibrated in a location with a similar climate (in Kimberly, southern ID). Furthermore, the Kimberly Penman equation is the model used for the USBR Pacific Northwest Region's AgriMet crop water use program [Palmer, 2008].

3. Datasets

3.1 Overview

When an end-user looks at a forecast ET_{rc} for any given point in space, the information contained in the single-value forecast ET_{rc} is enhanced by comparing it to the historical context provided by a 30-year ET_{rc} climatology. The forecast dataset used is the National Digital Forecast Database (NDFD) (see Section 3.2), while the climatology data are derived from North American Land Data Assimilation (NLDAS) dataset (see Section 3.3).

Each forecast—daily or seven-day—is compared to the established long-term behavior of values, which here means the average, minimum, 90% exceedance, median (50%

exceedance), 10% exceedance, maximum, and variance of the values observed for the identical period of the year over the 30-year period 1980-2009.

Note that there are significant differences between the two datasets. The NLDAS data from which the climatology is derived is not necessarily unbiased with respect to the forecast data, so there may be some systemic error built into the comparison between the two. Further, the two datasets are at different spatial resolutions: the NDFD operates at a resolution of 2.5 km, while the native resolution for the NLDAS dataset is 12 km. Development of an unbiased climatological dataset at an identical spatial resolution is ongoing.

3.2 Forecast dataset: National Digital Forecast Dataset (NDFD)

The NWS NDFD consists of gridded forecasts of sensible weather elements created by National Weather Service Forecast Offices. The four NWS forecasted weather elements used to create FRET are as follows: sky cover, wind speed U_{κ} , surface temperature T , and dewpoint temperature T_d or Relative Humidity. Further details on these datasets follow:

Sky cover:

- the expected amount of opaque clouds (in percent) covering the sky,

Wind speed U_{κ} :

- sustained 10-meter sustained wind speed (in knots),

Temperature T :

- expected 2-meter temperature (in degrees Fahrenheit).

Dewpoint temperature T_d or Relative Humidity:

- expected 2-meter dewpoint temperature (or relative humidity).

Additional information about NDFD grids and weather elements is available online at <http://www.weather.gov/ndfd/>

3.3 Climatological dataset: North American Land Data Assimilation System (NLDAS)

The following section presents the development of the climatology of Penman-Monteith and Kimberly Penman ET_{rc} developed from the North American Land Data Assimilation System (NLDAS) dataset. NLDAS data are hourly, available from January 1, 1979, to within days of the present, at a 0.125-degree spatial resolution across CONUS. Of the 11 NLDAS variables available [NLDAS, 2010] the complete Jan 1, 1980 – Dec 31, 2009 time-series of hourly grids of the following variables were used:

Surface radiation, R_{in} and L_{in} :

- DSWR, Surface Downward Shortwave Radiation Flux [W/m^2],
- DLWR, Surface Downward Longwave Radiation Flux [W/m^2].

Here R_{in} and L_{in} (shortwave and longwave downward radiation, respectively) are explicit definitions of the radiative fluxes that for the NDFD are estimated from CC_{daily} in Equation (9).

Wind speed, U_2

- UGRD, 10-meter U wind [m/sec],
- VGRD, 10-meter V wind [m/sec].

Surface wind speed data are stored in the form of UGRD10m and VGRD10m, two orthogonal horizontal vector components of the hourly mean wind velocity [m/sec] at 10 m. These data are converted to the required scalar wind speeds at 2 m by accounting for the vertical wind speed profile, as follows:

$$U_2 = \sqrt{U_{10_x}^2 + U_{10_y}^2} \left(\frac{2}{10} \right)^{\frac{1}{7}}, \quad (21)$$

where the square-root term represents the scalar magnitude of the sum of the x and y (or $UGRD$ and $VGRD$) wind vectors, and in the second term, the exponent scales the wind speed from the elevation of the denominator to the elevation of the numerator in meters, i.e., in this case speeds at 10 m to speeds at 2 m.

Temperature, T

- TEMP, 2-meter Temperature [K].

Humidity, T_{den}

- SPFH, 2-meter Specific Humidity [kg/kg].

Station pressure, Pa

- PRES, Surface Pressure [Pa].

Daily driving data are then derived from these hourly inputs.

To describe the time-constant geography of the underlying grid, the following fixed grids are necessary:

- Elevation [m],
- Latitude [deg],
- Longitude [deg],
- Land mask.

VARIABLES

Symbol	Variable	Units
a	calibrated parameter in $f(U_2)$	$\text{W/m}^2/\text{Pa}$
a_{KP}	seasonal parameter in Kimberly Penman vapor transfer function	$\text{W/m}^2/\text{Pa}$
a_s	Angström coefficient	-
a_{sa}	calibrated parameter in Savinov-Angström equation	-
a_v	ratio of water vapor eddy diffusivity::eddy viscosity in neutral conditions	-
b	calibrated parameter in $f(U_2)$	$\text{J/m}^3/\text{Pa}$
b_{KP}	seasonal parameter in Kimberly Penman vapor transfer function	$\text{J/m}^3/\text{Pa}$
b_s	Angström coefficient	-
c_p	specific heat capacity of moist air	J/kg/K
c_s	volumetric heat capacity of the soil	$\text{J/m}^3/\text{K}$
c_w	specific heat capacity of water	J/kg/K
d	zero-plane displacement height	m
d_r	relative earth-sun distance	-
e_a	actual vapor pressure	Pa
e_{sat}	saturated vapor pressure	Pa
f_{dir}	direct beam fraction of R_m	-
$f(U_{\tilde{\kappa}})$	vapor transfer function for $U_{\tilde{\kappa}}$	$\text{W/m}^2/\text{Pa}$
$f(U_2)$	vapor transfer function for U_2	$\text{W/m}^2/\text{Pa}$
h	crop height	m
i	hour of day	-
k	von Kármán constant	-
n	number of observations per day	-
n_p	days past perihelion	-
n_s	observed sunshine hours	hours
r_a	aerodynamic resistance	sec/m
r_s	stomatal resistance	sec/m
t	time	sec
t_{sr}	sunrise time, relative to solar noon	hours
t_{ss}	sunset time, relative to solar noon	hours
$\tilde{\kappa}_m$	height at which $U_{\tilde{\kappa}}$ is measured	m
$\tilde{\kappa}_v$	height at which e_a is measured	m
$\tilde{\kappa}_{0m}$	roughness height for momentum	m
$\tilde{\kappa}_{0v}$	roughness height for water vapor	m
A_d	heat flux advected from control volume	W/m^2
C	biochemical heat flux and storage change	W/m^2
CC	cloud cover	%
CC_{daily}	mean daylight cloud cover	%
DTR	diurnal temperature range	K
E_A	drying power of the air	W/m^2
$E[n]$	climatological daily number of daylight hours for month of interest	hours
ET	actual evapotranspiration	W/m^2
ET_{rc}	reference crop evapotranspiration	W/m^2
G	ground heat flux	W/m^2
H	sensible heat flux	W/m^2

Symbol	Variable	Units
I	hourly weighting index for cloud cover	-
J	Julian day of year	-
L_{in}	downward longwave radiation	W/m ²
L_n	net longwave radiation	W/m ²
L_{out}	upward longwave radiation	W/m ²
N	maximum daily sunshine hours	hours
Prp	precipitation	mm/day
P_{atm}	atmospheric pressure	Pa
P_{rad}	pan radiation factor	-
Q_n	net available energy for evaporation	W/m ²
R_{cs}	clear sky shortwave radiation	W/m ²
R_{in}	downward shortwave radiation	W/m ²
R_n	net shortwave radiation	W/m ²
R_{toa}	top-of-atmosphere or extra-terrestrial shortwave radiation	W/m ²
T	air temperature	K
T_{dew}	dewpoint temperature	K
T_i	i^{th} air temperature observation	K
T_i	inflow temperature	K
T_{max}	maximum air temperature	K
T_{min}	minimum air temperature	K
T_o	outflow temperature	K
T_s	surface temperature	K
T_p	precipitation temperature	K
$U_{\tilde{z}}$	wind speed at height \tilde{z} m	m/sec
U_2	wind speed at height 2 m	m/sec
VPD	vapor pressure deficit	Pa
W	heat storage in evaporating layer	J/m ²
Z	elevation	m
α	albedo	-
γ	psychrometric constant	Pa/K
δ	declination	rads
ε	emissivity	-
ε_{in}	inward (atmospheric) emissivity	-
ε_{out}	outward (surface) emissivity	-
ε'	net emissivity	-
λ	latent heat of vaporization	J/kg
π	pi	-
Q_a	moist air density	kg/m ³
Q_w	water density	kg/m ³
σ	Stefan-Boltzmann constant	W/m ² /K ⁴
φ	latitude	rads
Δ	slope of the saturated vapor pressure curve	Pa/K
Δt	period over which G is estimated	secs
$\Delta \tilde{z}$	effective soil depth	m
Θ	moisture availability for evaporation	[-]
Λ	longitude	rads

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